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RESEARCH NOTE

A NOTE ON THE GOOCHLAND COUNTY, VIRGINIA, EARTHQUAKE OF MARCH 15, 1991

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ABSTRACT

The Goochland county, Virginia, earthquake of March 15, 1991 ($m_b(Lg)$=3.8), was the largest earthquake recorded in the central Virginia seismic zone (CVSZ) since the Cunningham, Virginia, earthquake of 1984 ($m_b(Lg)$=4.2). The 1991 event was felt over 23,000 km^2 with a maximum MM V epicentral intensity. The preferred depth of focus was 12.5 km as determined graphically from the T^2 vs X^2 plot, whereas the computer program HYPOELLIPSE gave a more model-sensitive depth of 15 km. The average focal depth for shocks in this zone is 8.6 km, and the 90% quantile depth is 13.3 km. This earthquake is important because it is the largest shock that has occurred near the base of the CVSZ since network recording begun in 1978.

P wave first motions and $(S_V/P)_Z$ amplitude ratios define a focal mechanism exhibiting primarily strike-slip faulting with a north-south or east-west strike. The P axis trends northwest, similar to deeper focus earthquakes (>8 km) in the CVSZ. P wave spectral analyses indicate a corner frequency at 8 Hz and a low stress drop level of under 100 bars.

INTRODUCTION

On March 15, 1991, at 1:54 (EDT), the central Virginia area from Richmond to Staunton was shaken by a magnitude $m_b(Lg)$=3.8 earthquake. The earthquake was located in Goochland County, 60 kilometers southeast of Charlottesville and 56 kilometers northwest of Richmond, near the hamlet of Sandy Hook at a depth of 12.5 to 15 km. The earthquake had maximum intensity MM V and total felt area approximately 23,000 square kilometers.

Three associated seismic events were recorded: one foreshock at 20:40 (EDT) of March 14 (M_D=1.3) and two aftershocks, at 2:05 and 16:05 of March 15 (M_D=0.4; 0.1). Expectably, none of the associated shocks were felt.

The March 15 earthquake occurred in the Central Virginia Seismic Zone (CVSZ), in the same general location as the largest shock within the zone, with maximum intensity MM VII and m_b=5, on December 22, 1875 (Oaks and Bollinger, 1986). The CVSZ has exhibited low-level, persistent seismicity for more than two centuries (Bollinger and Sibol, 1985). The earthquake of March 15, 1991, is the largest earthquake recorded in the CVSZ since the Cunningham earthquake (near Charlottesville) of August 17, 1984 with $m_b(Lg)$=4.2 (Davison et al., 1984). The hypocentral depth of the March 15 earthquake, near the base of the brittle-ductile transition as defined by the base of the seismicity, makes it an important event in the study of the zone. The following presents an analysis of its hypocenter location, intensity level and distribution, focal mechanism, and P wave spectral content.

HYPOCENTER LOCATION

Arrival-time recordings from the seven stations of the Virginia Tech central Virginia subnetwork were used for the location of the earthquake of March 15 (figure 1), its foreshock and the two aftershocks. The location parameters of the four events are summarized in Table 1, as calculated from the location program HYPOELLIPSE (Lahr, 1980). The magnitude of the

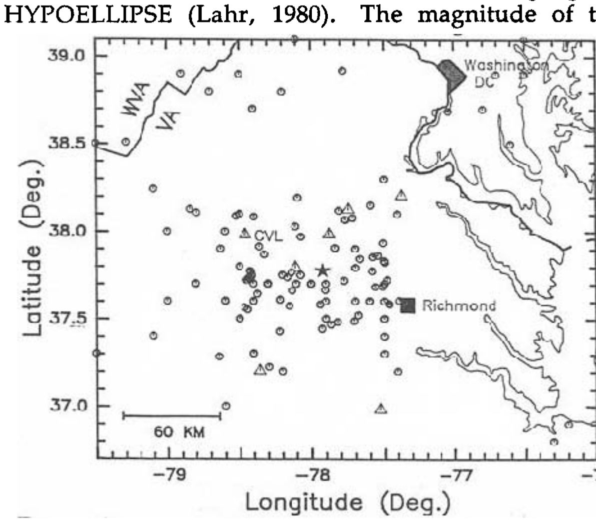

Fig. 1. Epicenter (1774-1991) map for the Central Virginia Seismic Zone and for the earthquake of March 15 (star). Also shown are the recording stations (open triangles) used to determine the hypocenter location. Call letters indicated only for the Charlottesville (CVL) station.

Table 1
HYPOELLIPSE Output of the Earthquakes of March 15, 1991

UTC	Lat N	Lon W	Depth (km±ERZ)	No Sta	MAG M_D	GAP (°)	RMS (sec)	ERH (km)	DMIN (km)
F 0140	37°44.05′	77°53.93′	12.1±3.8	6	1.3	198	0.30	1.8	20
M 0654	37°44.71′	77°54.63′	15.0±1.7	7	3.8*	116	0.21	0.7	18
A 0705	37°44.29′	77°54.55′	10.4±4.5	6	0.4	196	0.30	1.2	19
A 2105	37°46.32′	77°55.83′	14.7±2.7	4	0.1	257	0.30	4.8	16

M main event, F foreshock, A aftershock, * $m_b(Lg)$

main event was $m_b(Lg)$=3.8, calculated from the WWSSN station BLA seismogram.

The depth distribution of instrumentally located hypocenters from 1977 to 1990 for the CVSZ is shown in Figure 2. The deeper than usual depth of the March 15 earthquake for the CVSZ motivated this study. Because the mainshock S wave energy saturated most analog and digital records, S wave arrival times were difficult to pick, especially at the stations nearest to the epicenter. To obtain better S-P arrival time estimates for the main event, the unclipped foreshock and aftershock waveform recordings were compared, cycle by cycle, with the main event. Each station's recording of the foreshock and aftershocks was correlated with that same station's main event recording. First, the lag time (time offset between the two correlated time series) was determined from the correlations for each pair of recordings (mainshock with either foreshock or one of aftershocks), and that lag was used to time shift and align the seismic recordings to allow for a direct cycle-to-cycle waveform comparison. The resulting waveform comparisons showed that the main shock traces correlated peak-to-peak and trough-to-trough with the foreshock (except, of course, for the amplitude of the signals) at all recording stations. The digital time series from station NA2 are shown in Figure 3 as a sample of these comparisons. The two aftershock waveforms also agreed generally with the main event, but were not a perfect match. Consequently, S-P arrival time estimates of the foreshock for the Goochland (GHV) and Charlottesville (CVL) stations were used for the main event's hypocentral parameter determination. For the remaining five recording stations, the S-P arrival time estimates of the foreshock were compared to the ones picked of the main event recordings, and were found to be in good agreement.

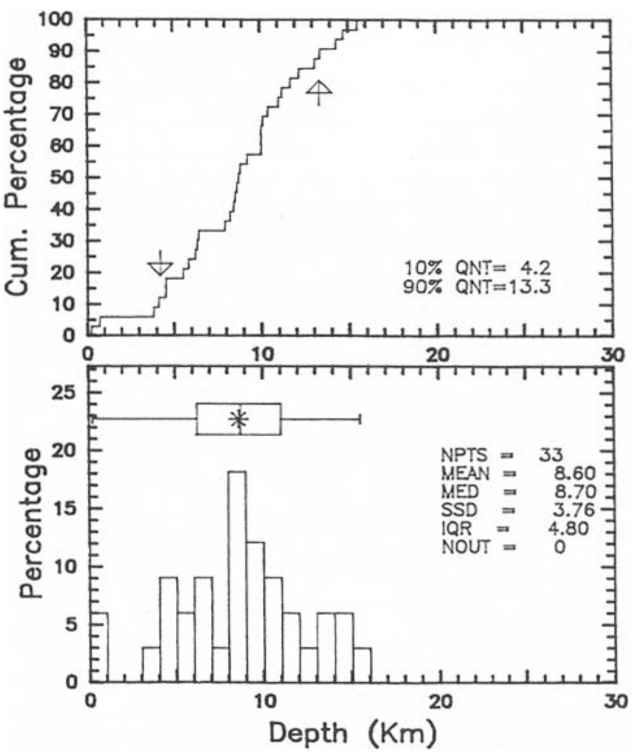

Fig. 2. Earthquake depth distribution in the Central Virginia Seismic Zone for 33 events (NPTS) instrumentally located, from 1977 to 1990, with ERZ ≤ 5 km. The MEAN depth for the seismic zone is 8.6 km, with a sample standard deviation 3.8 (SSD), interquartile range 4.8 (IQR) and 0 outliers (NOUT). The 90% quantile (QNT) depth is 13.3 km.

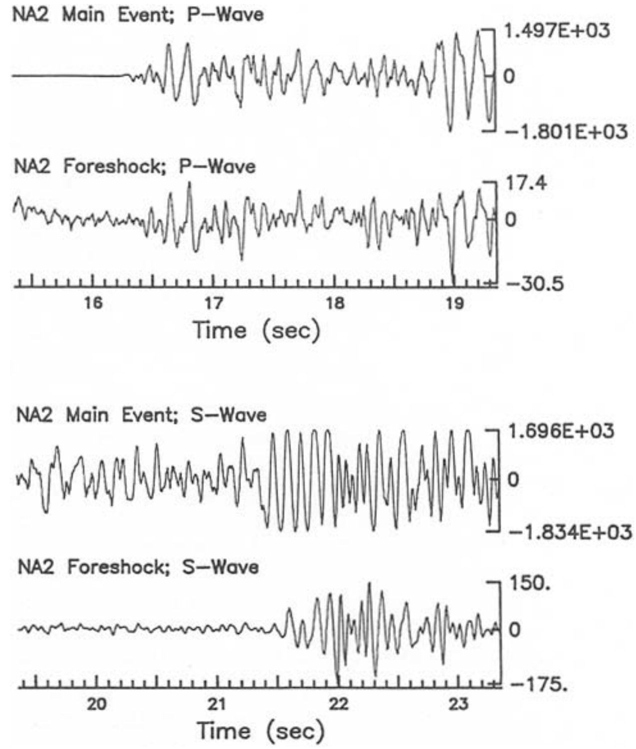

Fig. 3. Waveform comparison of the March 15 mainshock and its foreshock. Digital recordings at 100 samples per second, of the station NA2 vertical component, at 44.7 km from the epicenter. The vertical (amplitude) scales correspond to digital counts.

The observed waveform agreement between the foreshock and the mainshock suggests that both events occurred at or near the same location and with similar mechanisms. The hypocentral parameters calculated by HYPOELLIPSE (Table 1) for the main event and the foreshock, were found not to be in exact agreement and the depth estimates were different by approximately 3 km. Using the foreshock arrival times with the hypocentral parameters fixed at the values of the

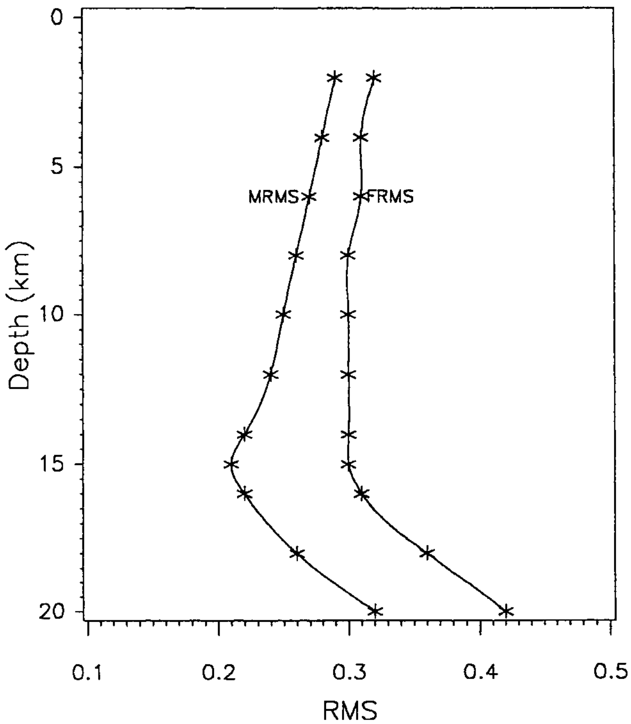

Fig. 4. RMS values calculated by HYPOELLIPSE, for fixed depths and the remaining hypocentral parameters free. MRMS = Main event RMS; FRMS = Foreshock RMS.

main event, HYPOELLIPSE yielded the same RMS statistics as the runs of the foreshock with free hypocentral parameters. Furthermore, Figure 4 demonstrates that HYPOELLIPSE could pick as a depth estimate for the foreshock any value between 8 and 15 km. However, for the main event, the depth solution of 15 km is more constrained, at least with the input data and the crustal model employed. Thus, both the foreshock and mainshock probably occurred near the same depth.

The crustal model, for the central Virginia area, input to HYPOELLIPSE consists of two layers over a half space the upper one of which has V_P=6.09 km/sec, V_P/V_S =1.73 and thickness 15 km (Bollinger et al., 1980). Thus, the mainshock hypocenter is located at the base of the first layer. An independent method to test the HYPOELLIPSE calculations of the origin time (OT) and focal depth, along with the input crustal model parameters V_P/V_S and average V_P, is the Wadati method (Wadati, 1933) and the T^2 vs X^2 method (see, e.g., Chapman and Bollinger, 1984).

In a Wadati plot S-P interval times versus P wave arrival times are assumed to have a linear relationship (figure 5). The intercept of the straight line with the abscissa is an estimate of the origin time (OT=06:54:08.9) and the slope of the line (0.72) is related to the ratio of the P and S wave velocities by the expression *slope*=V_P/V_S-1.0, which gives a V_P/V_S=1.72. Both OT and V_P/V_S estimates of the Wadati plot are in agreement with the OT of 06:54:08.3 calculated by HYPOELLIPSE and the velocity model ratio of the first layer of 1.73.

The T^2 vs X^2 data are plotted in Figure 6, where T is the P wave travel time (T=P arrival - OT from Wadati plot) and X is the epicentral distance. T, X, and V_P are related by the expression: $T^2=h^2/V_P^2+X^2/V_P^2$ where h is the hypocentral depth. Thus, in Figure 6, the slope of the line is related to the average P wave velocity above the focus by the expression *slope*=$1/V_P^2$, giving an average V_P=6.35 km/sec and the intercept with the ordinate $T^2 = 3.9 \pm 0.3$ yields a depth $h = 12.5 \pm 0.5 km$. Both estimates are in only fair agreement with the HYPOELLIPSE estimate of depth $h = 15 \pm 1.7$ km and the average P wave velocity used in the model of 6.09 km/sec for the first layer (0 to 15 km depth) and 6.5 km/sec for the second layer (15 to 36 km depth), since the analytical method had the seismic rays propagating to the more distant stations through the higher speed second layer.

In comparing the two hypocenter depth estimations, 15.0 km from the iterative algorithm HYPOELLIPSE, and 12.5 km from the Wadati T^2 vs X^2 method, the latter one is probably a better estimation of the true hypocenter, since it is relatively independent of the crustal model. The HYPOELLIPSE solution is highly dependent upon the thickness of the upper layer of the assumed velocity model. Finally, the HYPOELLIPSE calculations with V_P and OT values fixed at the values determined from the Wadati plot and the T^2 vs X^2 plot yield a depth estimate of 12.1 ± 0.3 km, in virtual agreement with the depth determined from the Wadati T^2 vs X^2 method.

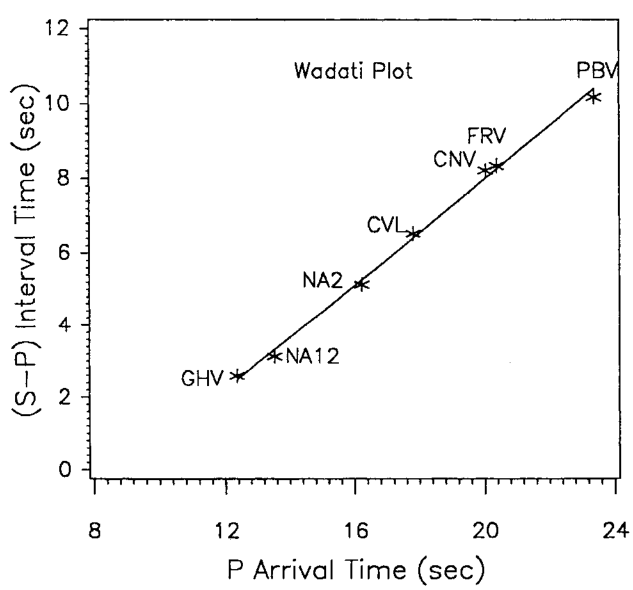

Fig. 5. Wadati Plot with main event S-P interval times plotted versus the P wave arrival time. A least squares straight line is fit to the data, with a resulting slope 0.72 and intercept of 8.9 sec. The slope indicates a V_P/V_S=1.72, and the intercept indicates an origin time of 06:54:08:9 (correlation coefficient: R^2=0.99).

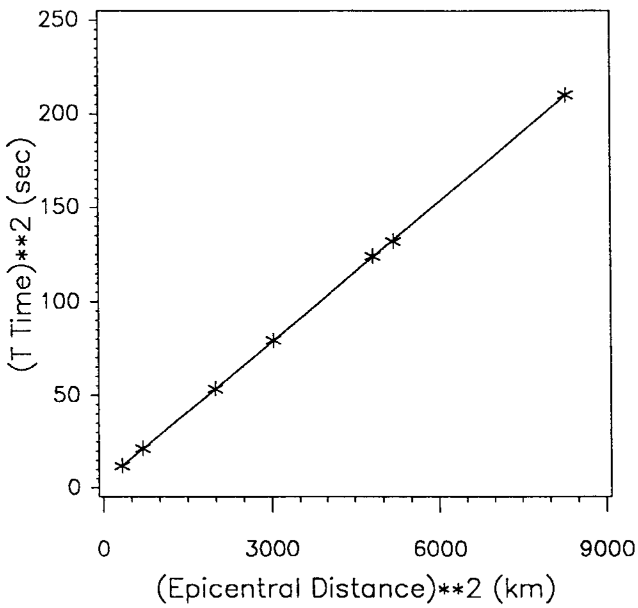

Fig. 6. T^2 vs X^2 plot, where T is the P wave travel time and X the epicentral distance. A least squares straight line is fit to the data, with a resulting slope 0.025, an intercept 3.9±0.3 and a resulting depth h=12.45 km with an average V_P=6.35 km/sec (correlation coefficient: R^2=1).

INTENSITY ANALYSIS

A total of 60 reports of the March 15 earthquake's effects were evaluated in an intensity survey of the area. Most reports were letter replies from residents in the area to requests for such information issued by the Virginia Tech Seismological Observatory and printed in local newspapers. Additionally, several intensity questionnaires, distributed two days after the earthquake in the epicentral area, were also received. A reconnaissance survey of the epicentral area conducted by Virginia Power (C. M. Robinson, 1991, written communication) included interviews with people that live in the area and an investigation of the area for possible damage (none was found). All the available reports were assigned intensity values according to the Modified Mercalli intensity scale, and are plotted in Figure 7. The results of this intensity survey are in general agreement with the intensity values presented in USGS PDE No 11-91, April 5, 1991, at those localities where both estimates were made.

The area of maximum intensity was at the MM V level, and included the instrumental location of the epicenter. Multiple reports from the same location (Goochland, Sandy Hook, Gum Spring, Oilville) enabled the determination of the meisoseismal area with some confidence, even though most residents were asleep at the time of the earthquake. Reports from local residents describe being awakened by a frightening, explosive sound, shaking of the entire house, rattling of the windows, dishes and doors, pictures swung out of place, and in one instance, kitchen cabinet doors jarred opened. Expectably, there were no damages reported from the earthquake of March 15. The investigation by Virginia Power in the meisoseismal area for possible damage, included observations of old and fragile brick chimneys and balanced rock monuments in a cemetery that showed no evidence of recent movement.

Several reports from the city of Richmond indicated that the central and northern section of the city apparently experienced more shaking than the southern section. In both areas, people were awakened by a strong, rumbling noise, and often some shaking. In some reports, the earthquake effects are described as those that would be caused by a heavily loaded truck or a railroad train passing by the house. The earthquake was felt over an area of approximately 23,000 km^2. This is at best a rough estimate, since at the time of the earthquake occurrence (1:54 AM EDT) most central Virginia residents were asleep and were not awakened by the sounds or the shaking.

An estimate of magnitude using a model from Sibol et al. (1987) method for eastern North America

$$m_b = 2.05 + 0.0219 I_0^2 + 0.0666 \log^2(FA),$$

where

$$IV \leq I_0 \leq VIII \text{ and } 3.0 \leq \log(FA) \leq 6.5,$$

where I_0 is the maximum Modified Mercalli intensity and FA is the felt area in square kilometers, yielded m_b=3.86 which is in good agreement with the observed $m_b(Lg)$=3.8 of the event.

FOCAL MECHANISMS

Because local recordings of the mainshock were clipped, (S/P) amplitude ratios could not be obtained for the focal mechanism solutions. However, the waveform agreement between the foreshock and the

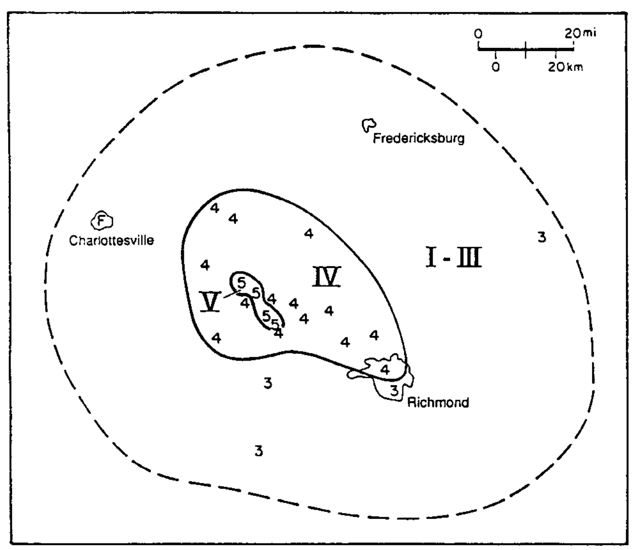

Fig. 7. Isoseismal Map of the March 15, 1991 CVA earthquake, with a total felt area of 23,000 km^2, IV intensity area of 2,500 km^2, and V intensity area of 100 km^2. F indicates areas where the earthquake was felt; Arabic numerals indicate MMI assigned to localities and Roman numerals correspond to isoseismal levels.

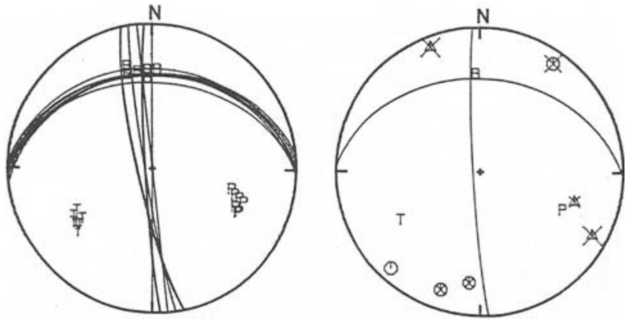

Fig. 8. Focal Mechanism Solutions of the March 15, 1991 CVA earthquake. Left: Lower hemisphere plots of the family of seven focal mechanism solutions along with their P, T and B axes. Right: Plot of the preferred solution and input data (circles=compressions; triangles=dilatations; X's=relative $(S_V/P)_Z$ amplitude ratios).

mainshock suggests that both events had similar mechanisms. Hence, a focal mechanism solution for the foreshock of March 15 at 01:40:23.04 UTC was obtained.

The focal mechanism solution was based on data consisting of seven P wave first motions and $(S_V/P)_Z$ amplitude ratio data from five unclipped station readings. These data were used as input for the computer program FOCMEC (Snoke et al., 1985). On the left of Figure 8, the family of seven solutions calculated from FOCMEC for 0 allowed P polarity and amplitude ratio errors (for a ratio of 2) are shown, along with the corresponding (T) tension, (P) pressure and (B) null axes. On the right of Figure 8, the preferred focal mechanism solution along with the input data are shown. The preferred solution was selected on the basis of sum of the squared error for all amplitudes (0.103). The focal mechanism solution parameters are summarized in Table 2 and indicate primarily strike-slip motion with a small component of normal faulting, on a north-south or east-west striking plane, dipping 86°. to the west and 35°. to the north, respectively. The focal mechanism solution of the foreshock (and probably also the mainshock) of the March 15 earthquake is similar to the focal mechanisms observed for the deeper earthquakes (>8 km) in the CVSZ by Munsey and Bollinger (1985); a mixture of dip-slip (reverse) and strike-slip faulting, and northwest trending P axes. The slight normal motion indicated for the March 15 shock is unusual.

Table 2

Preferred Focal Mechanism of the March 15, 1991 foreshock

Nodal Plane	Strike	Dip	Rake
1	N03°W	86°SW	-55°
2	N87°W	35°NE	-173°
Axis	**Trend**	**Plunge**	
P	S62°E	39°	
T	S58°W	32°	
B	N06°W	35°	

Sign convention on rake: 0° to -90° = left-lateral to normal faulting, -90° to -180° = normal faulting to right lateral

SPECTRAL ANALYSIS

The earthquake of March 15, 1991, generated a shear wave that saturated most stations in the VA TECH CVA subnetwork, except for NA12, which had a low signal to noise ratio. Therefore, at the seven recording stations of epicentral distances up to 100 km, only P wave spectra were examined. The first two seconds of the P wave recorded at station CNV (70 km epicentral distance), were used in the spectral analysis, and the spectrum corrected for the instrument displacement response is shown in Figure 9. Using the method developed by Snoke (1987), the mainshock P wave corner frequency (f_c) was determined at 8 Hz and the high frequency slope proportional to ω^{-2} (Figure 9). The gain calibration for station CNV is not well determined, therefore, the scaling of the ordinate axis in Figure 9 does not represent real displacement amplitude levels. For the constant stress model, P wave corner frequency f_c is related to stress drop $\Delta\sigma$ and seismic moment M_0 by

$$f_c = 4.9x\,10^6 v(\Delta\sigma/M_0)^{1/3}.$$

where v is phase velocity (α=6.35 km/sec), $\Delta\sigma$ in bars and M_0 in dyne-cm (Boore and Atkinson, 1987). Assuming that $m_b(Lg) \approx$ **M**, where **M** is moment magnitude, M_0 can be calculated from the expression **M** $= (2/3)\log M_0 - 10.7$ (Hanks and Kanamori, 1979, as in Boore and Atkinson, 1987). For the March 15, 1991, earthquake, $M_0 = 5.6x10^{21}$ dyne-cm and $\Delta\sigma$=95 bars. Note that the typical stress drop values reported for

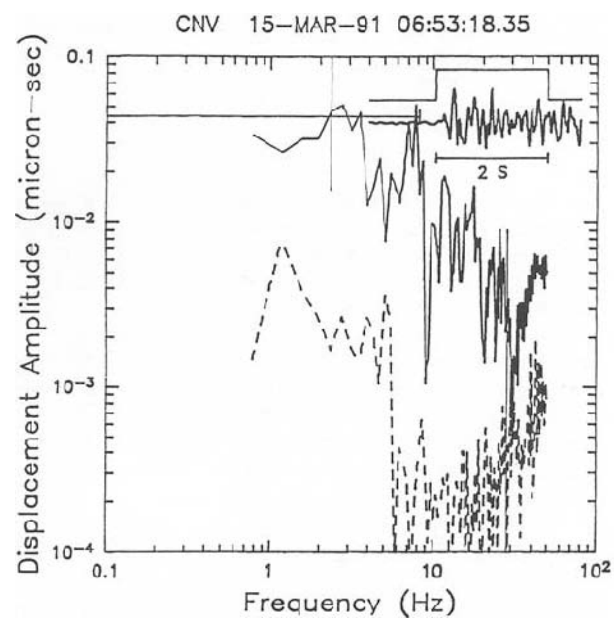

Fig. 9. Spectral Analysis for the P wave (80.03 to 82.03 sec) recorded at CNV (70 km epicentral distance). The corner frequency was determined at 8 Hz and the high frequency slope proportional to ω^{-2}. Vertical lines indicate the frequency bandwidth considered in the analysis, and the horizontal line corresponds to the chosen low-frequency spectral level. The pre-signal noise spectra are plotted as a dashed curve.

western North America are about 100 bars (Boore and Atkinson, 1987).

A second estimate of the stress drop is obtained from the amplitude spectrum of the Lg wave recording (15 seconds) at station BLA (Blacksburg, VA, 230 km epicentral distance) as shown in Figure 10. The spectrum of Figure 10 has been corrected for the instrument displacement response, and for anelastic attenuation (Chapman and Rogers, 1989). The estimate of the low-frequency displacement spectral level Ω_0 is 0.7 micron-sec and the Lg corner frequency is approximately 2 Hz. An independent estimate of moment magnitude (M_0) is obtained directly from Figure 10, with the expression relating the low-frequency displacement spectral level and moment magnitude for a Brune spectral model (Boore and Atkinson, 1987),

$$M_0 = \Omega_0 4\pi v^3 \rho \sqrt{RR_x}/C ,$$

where Ω_0=0.7 micron-sec, Lg phase velocity v=3.5 km/sec, R=230 km hypocentral distance, R_x=100 km a constant introduced to account for the geometric spreading of surface waves, crustal density ρ = 2.7gm/cm^3, and C = 0.89 a scaling factor accounting for radiation pattern correction, free surface correction and partition between horizontal components correction, as defined in equation (2) in Boore and Atkinson, 1987. The resulting estimate for M_0 is $1.7x10^{21}$ dyne-cm, yielding an **M** = 3.5. Using the Lg corner frequency of 2 Hz at BLA (Figure 10) and velocity v=3.5 km/sec yields a stress drop $\Delta\sigma$=3 bars. However, the Lg corner frequency may be underestimated at BLA due to inaccurate anelastic attenuation correction, because of the large epicentral distance. Assuming f_c=8 Hz and v=6.35 km/sec obtained from the P wave from CNV, at a much closer distance, yields $\Delta\sigma$ =30 bars.

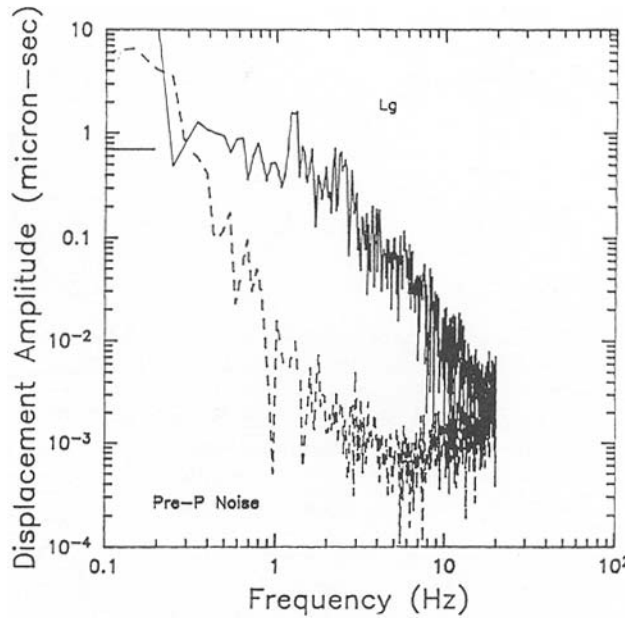

Fig. 10. Spectral Analysis for the Lg wave (15 sec) of the March 15 earthquake, recorded at BLA (230 km epicentral distance). The low-frequency displacement spectral level Ω_0 was chosen at 0.7 micron-sec as indicated by the horizontal line, and the corner frequency f_c at 2 Hz.

CONCLUSIONS

The earthquake of Goochland county, Virginia, of March 15, 1991, had a depth of 12.5-15 km, which puts it at the base of the seismogenic zone in the central Virginia. The earthquake was felt over 23,000 km^2, had a maximum MM V intensity, an $m_b(Lg)$ of 3.8, and $M_0 = 5.6x10^{21}$ dyne-cm. Focal mechanism solutions indicate strike-slip faulting and a NW trending P-axis; both of these characteristics are consistent with the deeper earthquakes (>8 km) in the zone. Spectral analysis shows a P wave corner frequency at 8 Hz, consistent with a stress drop of 3 to 95 bars.

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